Oxygen-18 variations in a global ocean model

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Abstract

The ratio of ^{18}O to ^{16}O in water of the world oceans is examined using the GISS ocean general circulation model. This ratio is governed by fractionation during evaporation and sea-ice formation and by the isotopic content of precipitation and runoff entering the oceans. The δ^{18}O$_w$ at any point depends on a balance between local evaporation, precipitation, and regional ocean and sea-ice processes. Surface δ^{18}O$_w$ equilibrates after 30–50 model years to produce a pattern of high values (0.5–1.5 ‰) in the sub-tropics (highest in the Indian and Atlantic oceans) and low values (-0.5–-1 ‰) in the mid-to-high latitudes. Arctic values are low especially where runoff enters from Arctic rivers. Comparisons with data show that regional linear relationships between δ^{18}O$_w$ and salinity are reasonably captured by the model. Surface data from the entire ocean show a clear distinction in slope between the tropical oceans and the extra-tropics, however, departures from the linear relationships are significant. Importantly for attempts at calculating paleo-salinities, the seasonal and longer-term temporal gradients ($\Delta\delta^{18}$O$_w/\Delta S$ at a fixed point) are not generally equal to the spatial gradient (at a fixed time).

Introduction

Much of the evidence for long term variability in the oceans comes from examination of the ratio of oxygen isotopes (^{18}O to ^{16}O) in $CaCO_3$ deposits in deep sea cores and, increasingly, from the skeletal remains of corals. These records faithfully preserve the ambient ratio of the oxygen isotopes in the surrounding seawater, modified by a temperature dependent fractionation that occurs as the carbonate precipitates from the seawater. The local temperature record is clearly of climatic significance, but the background ratio in the seawater is itself a function of the climate. It is modified by fractionation occurring as water evaporates or freezes, and of mixing with other water masses through precipitation, runoff, vertical mixing, convection and horizontal advection. Also, since ice sheets are built up from high latitude precipitation that is very depleted in ^{18}O, the growth and decay of glaciers leads to large changes in the average seawater oxygen isotope ratio. Hence, except in exceptional circumstances, it is difficult to unambiguously ascribe a single climatic cause to any event recorded in the carbonate.

If other proxy data (i.e. faunal assemblages, alkenone ratios or Sr/Ca ratios in corals) can be found to constrain the local temperature variations at a site, then the residual record from the carbonate record can be used to directly infer changes in $\delta^{18}O_w$[1]. Given the importance to paleoceanographic modelling of the surface salinity field it is tempting to use linear regressions based on the correlations that exist in today's oceans [*Craig and Gordon*, 1965; *Fairbanks et al.*, 1992] between $\delta^{18}O_w$ and salinity to estimate paleosalinity [*Duplessy et al.*, 1991; *Rostek et al.*, 1993]. However, the assumption that underlies these estimates is that the long term temporal gradient between $\delta^{18}O_w$ and salinity at a point is equal to the present day spatial gradient. This has yet to be demonstrated and, using quite conservative assumptions, it has been argued that the importance of advection and runoff in determining the gradient locally implies that this will not in general be the case [*Rohling and Bigg*, 1998].

Knowledge of the present day distribution of $\delta^{18}O_w$ comes mainly from the GEOSECS expeditions [*Ostlund et al.*, 1987] and other smaller scale surveys. Knowledge of the seasonal or longer-term variability of

$\delta^{18}O_w$ is even sparser, confined to a few time series off atolls in the tropical Pacific [*Wellington et al.*, 1996] and in the marginal N. Atlantic [*Strain and Tan*, 1993].

Previously, stable water isotopes in the present day climate have been successfully modelled with atmospheric general circulation models [*Jouzel et al.*, 1987] and attempts have been made to estimate surface $\delta^{18}O_w$ based on the modelled evaporative fluxes and precipitation [*Juillet-Leclerc et al.*, 1997]. Some ocean model experiments have included a highly idealized $\delta^{18}O_w$-like tracer [*Mikolajewicz*, 1996].

Ocean Model and Tracers

The ocean model is derived from the GISS coupled atmosphere-ocean general circulation model [*Russell et al.*, 1995]. It is a mass- and tracer-conserving primitive equation model with $4° \times 5°$ resolution incorporating a free ocean surface, sea-ice thermodynamics and advection, and a linear upstream scheme for advecting the tracers and their gradients. The model is initialised with climatological ocean temperature and salinity fields [*Levitus et al.*, 1994] and uses full flux boundary conditions at the sea surface using climatological surface atmospheric variables (incoming shortwave and longwave radiation, air temperature, relative humidity, pressure, precipitation, and winds). Outgoing longwave, latent and sensible heat fluxes and evaporation are calculated prognostically. Runoff is calculated by the land surface component of the coupled model consistent with the atmospheric climatologies. A small artificial feedback controls the absolute amount of evaporation to ensure long term mass balance.

Full flux boundary conditions are mandated by the need to separate evaporation and precipitation in order to calculate the isotopic exchange at the surface but, over the long term (> 100 years), this can lead to significant drift in the model. This tendency, combined with computing constraints, preclude running the model to equilibrium. Hence, this study is mainly concerned with the dynamics of the surface layers while the deep ocean remains much as it was initialised.

The model carries the total mass of ^{18}O as a tracer and conserves it exactly. The field is initialised using basin latitude-depth fields derived from GEOSECS and recent Arctic data [*Bauch et al.*, 1995]. Any longitudinal structure in the resultant fields therefore derives from the model dynamics. The isotopic con-

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[1] The standard δ-notation with respect to the isotope ratio in Standard Mean Ocean Water (R_{SMOW}) for seawater isotopic values is used where $\delta^{18}O_w = ((^{18}O/^{16}O)/R_{SMOW} - 1) * 1000$.

tent of monthly precipitation is set from the regression formula $\delta^{18}O_P = -11.88 + 0.345T - 0.0022P$ [*Gat and Gonfiantini*, 1981] (where T and P are the monthly mean temperature ($^\circ$ C) and precipitation (mm)), based on the WMO precipitation monitoring station network (which unfortunately has poor coverage over high latitudes and limited coverage over the open ocean). The isotopic exchange at the ocean surface is calculated using a kinetic fractionation model [*Gat*, 1996]. Atmospheric water vapour is deemed to be in isotopic equilibrium with local precipitation. Condensation also occurs in isotopic equilibrium. Equilibrium fractionation at a change of phase is fully temperature dependent. Again, a feedback controls the amount of isotopic exchange with the atmosphere to ensure that the tracer mass is also balanced.

The upper ocean fields and tracers equilibrate after about 30-50 model years after which they generally remain stable. All subsequent comments refer to the model state for the years 50-59.

The model has a reasonable representation of the upper ocean temperature and salinity except in the North Atlantic region. The total winter sea ice mass is around 80% (N. Hemisphere) and 130% (S. Hemisphere) of climatological values. Model deficiencies include: too little runoff from tropical rivers (which are also too high in $\delta^{18}O_w$), a weak Gulf Stream (40 Sv off Florida at 30° N, about half of observed values), and stronger Kuroshio and Antarctic Circumpolar Currents than observed. The N. Atlantic overturning streamfunction maximum (20 Sv) is too shallow and is too far south.

The pattern of upper ocean $\delta^{18}O_w$ is largely in agreement with the GEOSECS data, (fig. 1) with high values (0.5–1 $^0/_{00}$) in the sub-tropics and low values (-0.5–-1 $^0/_{00}$) in the mid-to-high latitudes. Arctic values are low especially where there is a large amount of depleted runoff entering from Arctic rivers, although values are slightly higher than seen in data for the region (possibly due to the too heavy isotopic ratio of regional precipitation). The model deficiencies mentioned above lead to low values of $\delta^{18}O_w$ (compared to data) in the Atlantic, especially in the Greenland Sea region. The distribution of $\delta^{18}O_w$ is roughly zonal, but significant departures are apparent in the tropics and between ocean basins. The errors that would arise from assuming a purely latitudinal profile could be as much as 1 $^0/_{00}$ (equivalent to around 4° C in a paleo-temperature calculation).

Relationships between $\delta^{18}O_w$ and S

Since the processes that affect surface $\delta^{18}O_w$ are also the processes that affect surface salinities, it is not surprising that a good correlation exists between these two fields. A comparison between the (Yrs. 50-59 average) upper 250m model output (fig. 2) and the GEOSECS data for the Atlantic shows that the model can reasonably reproduce the $\Delta\delta^{18}O_w/\Delta$S slope of the GEOSECS regression line (model slope 0.33, GEOSECS slope 0.52) despite differences in the $\delta^{18}O_w$ and S values taken separately. Results for the Southern Ocean (model: 0.53, data: 0.58) and Pacific (model: 0.52, data: 0.49) are closer (not shown). The model values for these comparisons are taken from roughly the same locations as the data, but no attempt was made to correct for seasonal variation. The deviations and wider spread in the Atlantic are probably due to a combination of the model deficiencies in the N. Atlantic mentioned previously, and too heavy Arctic waters biasing slightly the low salinity end member. The Pacific ocean results, with no large Arctic input, correlate better. Other upper ocean regional slopes are also well represented: for instance in the Arabian Sea (data slope 0.28 [*Rostek et al.*, 1993]) the model slope is 0.24. In the N. E. and S. Central Pacific (160–120° W 0–58° N and 175° E to 150° W 20–64° S respectively) model slopes are 0.42 and 0.80 compared with slopes of 0.54 and 0.68 seen in data [*Craig and Gordon*, 1965].

Values from the entire surface ocean (fig. 3) show that although a large number of points from the extra-tropics lie clustered around the GEOSECS regression line, $\delta^{18}O_w$ in the tropical oceans is only weakly correlated with salinity. The tropics seem to loosely follow a line with a slope of 0.18 (c.f. a slope of 0.11 in the tropical Atlantic [*Craig and Gordon*, 1965]). The spread around the regression lines is significant: ±0.5 ppt (±1.5 ppt) for fixed $\delta^{18}O_w$ in the extra-tropics (tropics).

Seasonal and longer term change in $\Delta\delta^{18}O/\Delta S$

There is no reason to expect seasonal changes in the $\delta^{18}O_w$/S relationship to follow the spatial gradient. Differences in seasonal phasing between evaporation, precipitation, mixing and advection (each with a separate isotopic signature) imply that only rarely will salinity and $\delta^{18}O_w$ vary in phase. The seasonal cycles and longer term variability in the model are examined at three points (fig. 4), one just south of Ice-

fig. 2

fig. 3

fig. 4

land, one near the Galapagos Islands, and the other near Clipperton Atoll in the Pacific. The patterns in $\delta^{18}O_w$/S space vary widely depending on the balance of processes at each point. Similarly, the changes in the annual average values show that the longer term temporal gradients do not necessarily follow the spatial gradients either. For instance, the changes seen in the Icelandic basin arise due to the freshening of the N. Atlantic and slow decay of the overturning stream-function and represent an advective change of source region for that water mass. These changes are due to the drift of the ocean model and therefore should only be considered indicative of the actual variability. However, the seasonal changes and general distribution should be fairly robust.

What implications do these results have for attempts to calculate paleo-salinities? Firstly, the model results underline the differences found regionally in the gradients of the mixing lines and point to the need for more detailed data coverage. Secondly, the wide spread of points around these regional lines would seem to preclude their use in precisely estimating salinity changes, particularly in the tropics. And finally, these results cast further doubt on the assumption that the temporal and spatial gradients are equal, though there remains a possibility that longer ($\gg 10$ years) averaging periods would produce clearer results.

The untangling of the various factors involved in creating the deep sea core and coral $\delta^{18}O$ carbonate record relies on a better understanding of the long term $\delta^{18}O_w$ variability. Further efforts using $\delta^{18}O_w$ as an ocean tracer and forward modelling the signal that would be recorded in carbonates should enable researchers to map modelled climatic events to the core and coral records and allow for a better interpretation of those records.

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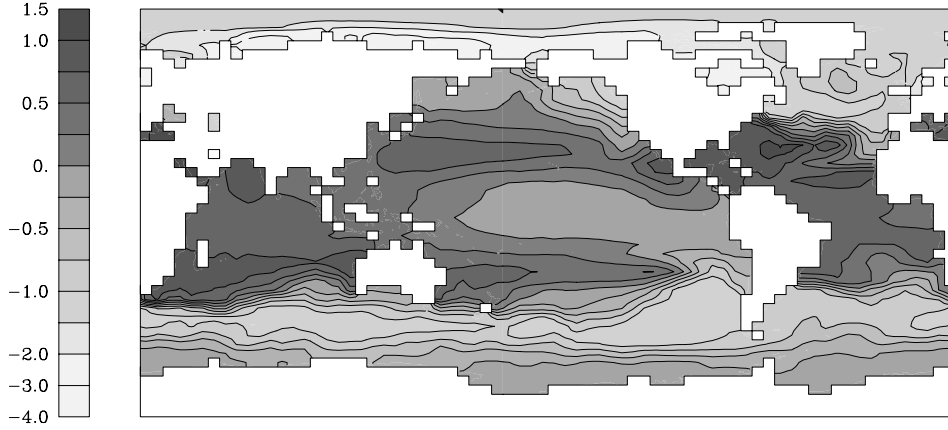
Figure 1. The averaged $\delta^{18}O_w$ (per mil) field in the surface ocean for model years 50-59. The pattern is mainly zonal, although significant meridional gradients occur. The largest departures from observations are the too low values occurring in the N. Atlantic.
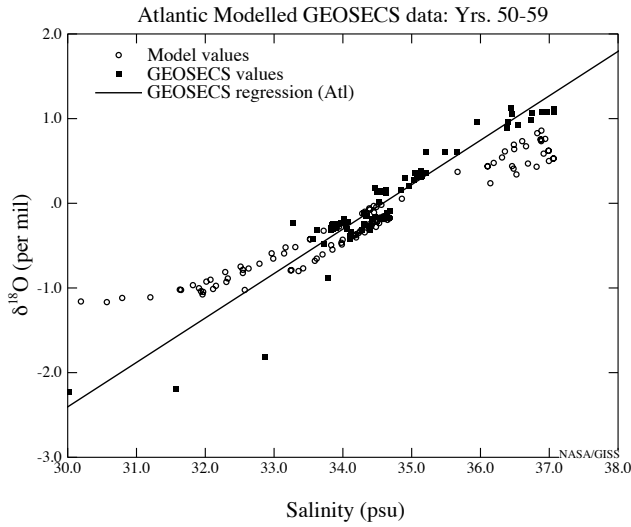

Figure 2. The relationship between $\delta^{18}O_w$ and S for GEOSECS data from the Atlantic and surface model values for Yrs. 50-59 taken from roughly the same points. The regression line shown is the fit to the GEOSECS Atlantic data.
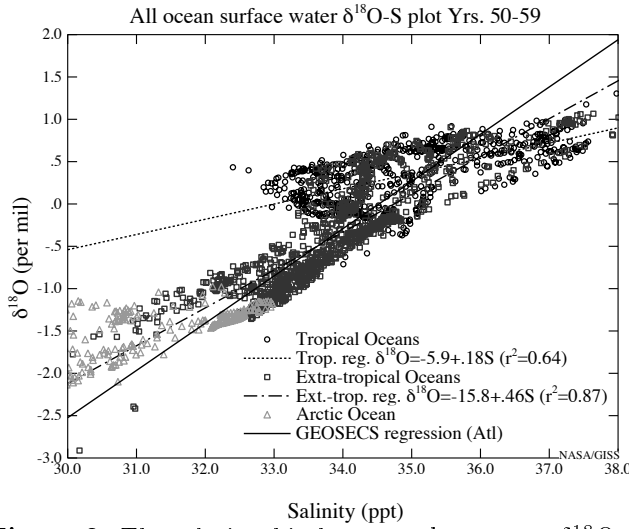
Figure 3. The relationship between the average $\delta^{18}O_w$ and S for all the surface points in the model for Yr. 50-59, split into the tropical, extra-tropical and Arctic Oceans. Although the regression line for extra-tropical waters lies near that for the GEOSECS data, large numbers of points (particularly those from the tropical oceans) follow a much less steep line where $\delta^{18}O_w$ is only loosely correlated with S.
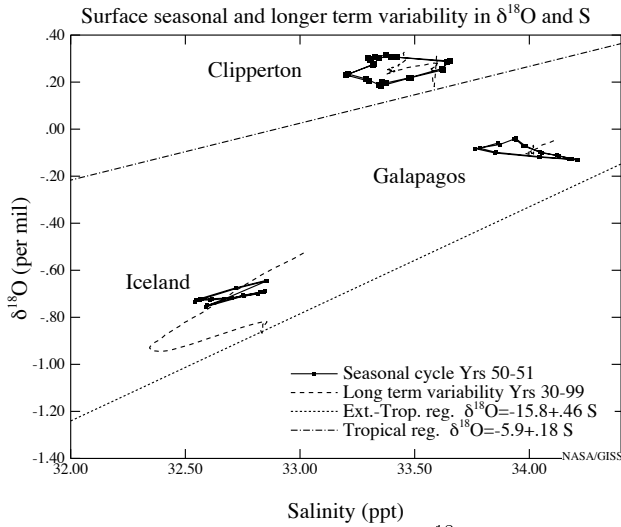

Figure 4. The relationship between $\delta^{18}O_w$ and S at three particular points in the model (a point south of Iceland, one near the Galapagos Islands and a point near Clipperton atoll in the Pacific), over a seasonal cycle and the longer term changes in the annual average. Note that the points do not necessarily lie on lines parallel to the spatial gradients